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The response of Black Rapids Glacier, Alaska, to the Denali earthquake rock avalanches

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ABSTRACT

We describe the impact of three simultaneous earthquake-triggered rock avalanches on the dynamics of Black Rapids Glacier, Alaska, using space-borne radar imagery and numerical modeling. We determined the velocities of the glacier before and after landslide deposition in 2002 using a combination of ERS 1/2 tandem, RADARSAT-1 and ALOS Palsar synthetic aperture radar data. Ice velocity above the debris-covered area of the glacier increased up to 14 percent after the earthquake, but then decreased 20 percent by 2005. Within the area of the debris sheets, mean glacier surface velocity increased 44 percent within two years of the landslides. At the downglacier end of the lowest landslide, where strong differential ablation produced a steep ice cliff, velocities increased by 109 percent over the same period. By 2007 ice velocity throughout the debris area had become more uniform, consistent with a constant ice flux resulting from drastically reduced ablation at the base of the debris. Without further analysis, we cannot prove that these changes resulted from the landslides, because Black Rapids Glacier displays large seasonal and interannual variations in velocity. However, a full-Stokes numerical ice-flow model of a simplified glacier geometry produced a reversal of the velocity gradient from compressional to extensional flow after five years, which supports our interpretation that the recent changes in the velocity field of the glacier are related to landslide-induced mass balance changes.

Keywords: InSAR, radar speckle tracking, Black Rapids Glacier, rock avalanche, glacier velocity, ice dynamics modeling
Large landslides can impact glaciers in many ways, notably by changing ice surface topography, glacier hydrology, albedo, stress state and ice velocity. Several authors have investigated the effects of large landslide debris sheets on glacier mass balance [e.g. Bull and Marangunic, 1967, 1968; Deline, 2005; Reznichenko et al., 2010; Vacco et al., 2010; Reznichenko et al., 2011] or on subglacial and englacial drainage [Reynolds, 2000; Gulley and Benn, 2007]. Others have examined the effect of changing ice surface elevation on hydraulic potential [Fountain and Vaughan, 1995; Fischer et al., 2005]. Few authors however, have attempted to quantify the effects of landslide debris on glacier flow, partly because pre-landslide velocity measurements are rarely available. Some researchers, however, have reported significant changes in glacier dynamics, including advances or surges, following large landslides [Tarr, 1910; Gardner and Hewitt, 1990; Deline, 2005].

Recent studies suggest that large landslides are more common in glacierized mountains than previously thought [Geertsema et al., 2006]. Ice velocity data for glaciers that have experienced rock avalanches, however, are rare. Bull and Marangunic [1968] attributed an ~10% increase in surface velocity of Sherman Glacier to the 1964 rock avalanche, based on a local velocity anomaly, but without pre-landslide measurements. During fieldwork in 2008, we observed that Sherman Glacier was advancing. Gardner and Hewitt [1990] attributed the 1987 and 1989 surges of Bualtar Glacier in Pakistan to three landslides in 1986, and M. Truffer (unpublished data, 2006; available at http://www.gi.alaska.edu/~truffer/) documented a 2006 surge of McGinnis Glacier, most of which was covered by debris of two rock avalanches triggered by the 2002 Denali earthquake. Shulmeister et al. [2009] provided a simple theoretical model for acceleration of a glacier following emplacement of a rock avalanche debris sheet, but were unable to validate the model with field data.

The objective of this study is to quantify and understand the response of Black Rapids Glacier to three rock avalanches triggered by the Denali earthquake in 2002. We compare pre- and post-landslide
ice velocities derived from space-borne radar satellite data and ground survey measurements, and use a full-Stokes ice-flow model to try to explain the observed changes in the velocity field.

2 SETTING

Black Rapids Glacier is a 40-km long, surge-type valley glacier in the eastern Alaska Range of interior Alaska (Figure 1). The glacier last surged in 1936-1937 [Hance, 1937] and has been extensively studied since then [Péwé, 1951; Post, 1960; Harrison et al., 1975; Heinrichs et al., 1996; Truffer et al., 1999; Rabus and Fatland, 2000; Truffer et al., 2001; Nolan, 2003; Amundson et al., 2006; Shugar et al., 2010; Shugar and Clague, in press].

The Mw7.9 Denali Fault earthquake, which occurred on 3 November 2002, triggered many landslides across the Alaska Range, including three large rock avalanches on Black Rapids Glacier (BRG-west, BRG-middle, and BRG-east; Figure 1). The landslides provide an opportunity to study the ice dynamics response to the instantaneous emplacement of ~30 x 10^6 m^3 of debris over ~11 km^2 of the glacier’s ablation zone [Shugar and Clague, in press]. Black Rapids Glacier, however, presents challenges for such a study, because of its spatially and temporally complex velocity field [Nolan, 2003] and difficulties in applying radar remote sensing methods to temperate glaciers [e.g. Massom and Lubin, 2006].

Since it was first surveyed in 1973, the surface velocity of Black Rapids Glacier has oscillated, with a period of approximately 12 years [Heinrichs et al., 1996]. The pattern and timing of these oscillations are consistent at km 20 and km 14, suggesting strong longitudinal stress coupling in the middle reach of the glacier [Nolan, 2003]. Heinrichs et al. [1996] attributed these flow variations to changes in basal motion rather than ice deformation. The decadal-scale oscillations on the upper half of the glacier have been large – more than 50% in the case of the km 8 and km 14 survey stations [Truffer et al., 2005]. No significant decadal velocity cycles have been observed below km 20, although survey data on the lower half of the glacier are limited. The difference in flow between the upper and lower parts of the glacier
may be due to different basal conditions, or to an effect on stress coupling by the Loket tributary, which
joins the main trunk of Black Rapids Glacier near km 24 (Figure 1).

Black Rapids Glacier also exhibits strong seasonal velocity fluctuations, with a peak in June and
much lower and generally less variable velocities in winter (October – February) [Heinrichs et al., 1996;
Rabus and Fatland, 2000]. Survey stations that exhibit marked seasonal speed variations (km 14, 20, 26,
and 32) lie on or near the Denali Fault. Three of these sites (km 14, 20, and 32) have also experienced
large long-term changes in annual velocity and are located in the ablation area [Heinrichs et al., 1996]. It
is not presently known if or how the Denali Fault influences these stations.

3 DATA AND METHODS

We generated pre-landslide glacier velocity maps using interferometric synthetic aperture radar
(InSAR), and post-landslide velocity maps using SAR speckle tracking. We then compared the satellite
data with ground-based survey measurements made since the 1970s and numerical model results of
glacier dynamics to examine if changes observed with SAR are characteristic features of post-landslide
glacier flow. A network of survey stations was established along the centerline of the glacier in 1973
[Heinrichs et al., 1995], and each station was denoted by its distance from the head of the glacier; for
example, survey station “km 26” is located 26 km from the head of the glacier (Figure 1). Table 1
provides a summary of all satellite data used in the study.

Data limitations constrained our choice of analytical techniques. We were able to perform
conventional interferometric processing with pre-landslide SAR data, but post-landslide SAR data were
decorrelated (see below), especially over bare glacier ice. As a result, we used SAR speckle tracking for
post-landslide velocity calculations.

3.1 Interferometric velocity measurements

Interferograms were generated from ERS-1/2 tandem data. The topographic signal was removed
using the 30-m DEM of the 1995 surface of Black Rapids Glacier produced by Shugar et al. [2010].
Atmospheric errors were filtered manually by masking glacierized areas and creating a smooth atmospheric screen, which was then subtracted from the interferogram.

Our approach differs from the similar method used by Mohr et al. [2003], in that we do not assume surface-parallel flow, but rather use flow direction vectors constrained by flow features and valley walls [Lang et al., 2004]. We digitized about 520 flow direction vectors in ArcGIS using a geocoded 2000 Landsat image of the glacier. These discrete vectors represent long-term averages of flow. We checked their accuracy by plotting long-term survey velocity vectors, based on data from Heinrichs et al. [1996] and M. Truffer (personal communication, May 2011). The digitized vectors were then interpolated to a regular grid using a Delauney triangulation to produce flow direction vectors parallel to the glacier surface for every pixel of the glacier [see Rabus and Lang, 2000]. In this manner, we were able to compute complete fields of surface parallel and surface normal (emergence) velocity. An assumption in this analysis, however, is that flow is invariant between image acquisitions. In the present study, the images were acquired only days apart (e.g. Table 1), and so this assumption is most likely valid.

Instead of using the Cartesian coordinate system (X, Y, Z) [Joughin et al., 1998], we rotated the flow field into a system with axes oriented parallel (II), vertically perpendicular (⊥), and horizontally transverse (t) to the glacier flow direction. The perpendicular component of velocity is nearly vertical, because Black Rapids Glacier has a very low surface gradient. The velocity, v, and the direction vectors, \( \eta \), use this same notation:

\[
v = v_\parallel \eta_\parallel + v_\perp \eta_\perp + v_t \eta_t
\]  

Assuming the transverse velocity, \( v_t \), is zero, velocities in the satellite line-of-sight direction (LOS) for the ascending (A) and descending (D) passes, are:

\[
v^{(A)} = v \cdot \eta^{(A)}_{LOS} = v_\parallel (\eta_\parallel \cdot \eta^{(A)}_{LOS}) + v_\perp (\eta_\perp \cdot \eta^{(A)}_{LOS})
\]

\[
v^{(D)} = v \cdot \eta^{(D)}_{LOS} = v_\parallel (\eta_\parallel \cdot \eta^{(D)}_{LOS}) + v_\perp (\eta_\perp \cdot \eta^{(D)}_{LOS})
\]
The assumption that $v_\perp = 0$ neglects ice flow parallel to topographic contours. It should be borne in mind that contour lines do not always run perpendicular to valley walls, but are concave or convex depending on location on the glacier. Contour-parallel flow could result in a non-stationary effect but it is small on the scale of several ice thicknesses, and due to the low surface gradient of Black Rapids Glacier [e.g. Truffer et al., 2001; Shugar et al., 2010]. Solving for the surface parallel and normal (emergence) components, we obtain, respectively:

\[ V_H = \frac{v^{(A)}_H \left( \eta_\perp \cdot \eta_{LOS}^{(A)} \right) - v^{(D)}_H \left( \eta_\perp \cdot \eta_{LOS}^{(A)} \right)}{\left( \eta_H \cdot \eta_{LOS}^{(A)} \right) \left( \eta_\perp \cdot \eta_{LOS}^{(A)} \right) \left( \eta_H \cdot \eta_{LOS}^{(D)} \right) \left( \eta_\perp \cdot \eta_{LOS}^{(D)} \right)} \] (3a)

\[ V_\perp = \frac{v^{(D)}_H \left( \eta_\perp \cdot \eta_{LOS}^{(A)} \right) - v^{(A)}_H \left( \eta_\perp \cdot \eta_{LOS}^{(A)} \right)}{\left( \eta_H \cdot \eta_{LOS}^{(A)} \right) \left( \eta_\perp \cdot \eta_{LOS}^{(A)} \right) \left( \eta_H \cdot \eta_{LOS}^{(D)} \right) \left( \eta_\perp \cdot \eta_{LOS}^{(D)} \right)} \] (3b)

Two descending-pass C-band ($\lambda=0.056$ m) ERS-1/2 tandem images from 8 and 9 October 1995, and two ascending-pass images from 11 and 12 October 1995 were used in this study (Table 1). The period spanned by these tandem images is one day (24:00 hours).

Possible errors in the InSAR data are difficult to quantify. For displacement calculations, InSAR is much more accurate than speckle tracking, thus most of the error derives from upscaling the error of the one-day tandem acquisitions to an annual equivalent. Random errors were considerably reduced by averaging to the lower spatial resolution of the speckle tracking (420 m x 315 m, see below). Systematic errors from residual atmospheric phase will, however, be amplified by upscaling from one day to one year. For the present study it is sufficient to state that the error of the InSAR-derived velocity maps is smaller than that of the velocity maps derived from speckle tracking over annual periods (see below).

Local errors in ice flow direction derived from the digitization and interpolation of flow vectors are estimated to be generally less than 2 degrees, which corresponds to a maximum 2% change in the surface-parallel velocity. Corresponding errors in the surface-perpendicular velocity are augmented by the smaller values of the perpendicular velocity.
3.2 Speckle tracking velocity measurements

If local phase shift gradients are too large due to rapid ice motion (more than one fringe per pixel), or random due to redistribution of snow, surface melt, or both, the phase can become decorrelated. As a result, the long temporal baselines of most spaceborne SAR systems result in generally poor phase correlation over temperate glaciers. A complementary approach to InSAR that overcomes this problem is speckle tracking. This method generates registration offsets between two SAR images in both slant range (satellite line-of-sight) and azimuth (parallel-to-satellite orbit) to map displacements [Michel and Rignot, 1999; Gray et al., 2001; Werner et al., 2001; Strozzi et al., 2008]. SAR intensity images are Fourier transformed, yielding the cross-correlation peak in pixel units [Michel and Rignot, 1999], and the image offset. Vertical velocity is ignored with speckle tracking, but it is safe to assume that the vertical component of the velocity is much smaller than the horizontal component.

Although less accurate than InSAR, radar speckle tracking has several advantages. The offset values are unambiguous 2D velocity measurements and do not require phase unwrapping, which can be difficult in steep mountains [Shugar et al., 2010] or for areas of fast ice flow [Short and Gray, 2004]. When InSAR phase becomes decorrelated so does the speckle pattern between the images in each pair. However, speckle tracking may still find meaningful offsets by tracking macroscopic features (edges, ridges, crevasses) even in the complete absence of speckle correlation. In other words, there is a transition between speckle tracking and feature tracking.

The speckle tracking technique is accurate to ~0.05 pixels [Werner et al., 2001; Murray et al., 2002], which corresponds to approximately 40 cm for a pixel size of 8 m (RADARSAT-1 Fine Mode). This amount is equivalent to 24 days of flow at 1.7 cm d$^{-1}$ or a year of flow at 0.1 cm d$^{-1}$. In the present study, speckle tracking measurements on stationary features in the Delta River valley east of Black Rapids Glacier using annual Radarsat-1 image pairs gave an average displacement error of 0.2 cm d$^{-1}$, or ~0.1 pixels. In comparison, one-day repeat (ERS tandem) interferograms measure range displacements to a small fraction of the radar wavelength, down to 1 mm accuracy assuming prior compensation for the
atmospheric error. To directly compare the speckle tracking and InSAR results in this study, we spatially averaged all velocity data over a series of image subsets (chips), each equivalent to 420 x 315 m on the ground.

We computed velocity maps of Black Rapids Glacier in the vicinity of the landslides using 27 C-band ($\lambda=0.056$ m) RADARSAT-1 images acquired during the melt seasons of 2003 to 2007 and four L-band ($\lambda=0.236$ m) ALOS Palsar images from the melt season of 2007 (Table 1). We determined offsets for consecutive monthly scenes (e.g. 4 July and 28 July 2005) and annual pairs (e.g. 28 July 2005 and 23 July 2006). We were thus able to quantify the glacier’s seasonal velocity pattern, as well as the longer-term changes over the years following the earthquake and landslides.

3.3 Ice dynamics modeling

We used an ice-dynamics model employing the finite element method to understand the response of a glacier to rock avalanche debris cover. The model solves the full Stokes equations in the Lagrangian formulation, which contain longitudinal stress coupling; we used a Weertman-type sliding law enforced iteratively through a Dirichlet boundary condition. The model was constructed in Matlab employing the finite element method implementation of COMSOL Multiphysics. We use default P2-P1 elements for the velocity and linear elements for the pressure, and stabilized by streamline diffusion and crosswind diffusion [Comsol, 2008].

The DEM produced by Shugar et al. [2010] provides a 1995 surface elevation profile of Black Rapids Glacier along its centerline. However, values for ice thickness and thus bed elevation are published for only a few discrete locations. Gades [1998, see his Figure 4.6] produced an ice-thickness map by interpolating cross-glacier echo sounding profiles collected about every 2 km between about km 6.5 and km 20. Other ice-thickness data were acquired along transverse ground penetrating radar (GPR) profiles that are unevenly spaced between km 14 and the terminus, with a maximum of about 4.9 km between profiles [Heinrichs et al., 1995, their Figure 10]. Interpolation of point bed elevations gives a
bed profile with much less detail and accuracy than the surface profile. To produce a consistent pair of
bed and surface elevation profiles, and to establish the glacier geometry at the time of the earthquake, we
let the DEM surface evolve by running the full-Stokes model (see below) for several years, updating the
surface elevation after each year by adding the difference between emergence velocity and local surface
mass balance. The surface and bed of the two-dimensional flowband were then interpolated to an even
50-m spacing using a spline. Net balance data used in the model come from mass balance measurements
made over the past 40 years, and were scaled by the area-altitude distribution. This approach makes the
simplifying assumption that centerline balances extend to the glacier margins [Arendt et al., 2002; c.f.
Berthier et al., 2010]. The slope of the edge of the debris sheets was kept at or below the angle of repose
(32 degrees). An adaptive triangular mesh was initially generated with 25-m spacing at the upper and
lower boundaries. The mesh was then refined by a factor of approximately two with the “meshrefine”
routine in Comsol. Figure 2 shows the model geometry, mass balance curve and mesh used to model ice
flow with the finite element method.

Ice was treated as an incompressible non-Newtonian fluid with a rheology described by Glen’s flow
law, \( \dot{\epsilon} = A \tau^n \), with \( A = 2.4 \times 10^{-24} \text{ s}^{-1} \text{ Pa}^{-3} \) for temperate ice [Cuffey and Paterson, 2010] and \( n = 3 \). We
used the 2D finite element method for the initial vertical flowband described above, to investigate if
small changes in surface topography, slope, and mass balance are adequate to cause the velocity changes
observed with satellites. The model implements a Weertman-style basal sliding law that includes
effective water pressure as a Dirichlet boundary condition, \( v_B = \frac{k \tau_B^m}{P_{\text{eff}}} \), where \( k \) is a sliding parameter
\( (25 \text{ m a}^{-1} \text{ bar}^{-m}) \), \( \tau_B \) is the basal shear stress, and \( m \) is a positive constant (2) [Bindschadler, 1983]. The
value of \( m = 2 \) is within the range described by Bindschadler [1983] and provides a good fit to the SAR-
derived velocity data. The effective pressure, \( P_{\text{eff}} \), is the difference between the ice overburden pressure
and water pressure. Water pressure was calculated as a fraction of ice overburden pressure, based on the
piezometric surface described by Truffer et al. [2001, their Figure 8] of 55 m below the ice surface.
Lateral drag along valley walls is approximately taken into account via shape factor modification to the flow-law coefficient, $A$ [Heinrichs et al., 1996].

From the 2D finite element model, 1D curves of both surface-parallel and surface-perpendicular velocity can be calculated along the glacier surface. We verified the model against the Ice Sheet Model Intercomparison Project-Higher Order Models (ISMIP-HOM) benchmark using test data and experiments from Haut Glacier D’Arolla [Pattyn et al., 2008].

Because bottom topography is poorly known, basal motion can only be roughly modeled in the full-Stokes model [Amundson et al., 2006]. We made no attempt to model the inter-annual velocity fluctuations of Black Rapids Glacier, which are thought to be related to changes in basal conditions [Truffer and Harrison, 2006]. We also did not consider what is likely the complex influence of the Loket tributary on the flow of the main glacier trunk (Figure 1). We therefore do not expect our results to closely match observed velocities but rather we are more interested in recognizing characteristic spatio-temporal patterns of velocity and elevation change. By comparing modeled patterns of velocity change to measurements, we tried to understand whether the observed velocity fluctuations were mainly caused by the rock avalanches or, instead, were the result of the quiescent evolution of this surge-type glacier.

We ran two experiments to investigate the role of landslide debris on glacier velocity. In our control run, we allowed the glacier surface and velocity structure to evolve for five years (2002-2007) without any surface debris, thus simulating the evolution of the glacier without a landslide. Landslide deposition was then separately simulated by locally increasing $h$ by the equivalent of a 2-m sheet of rock debris (bulk density 2400 kg m$^{-3}$) over a horizontal distance of 7 km, between km 25 and km 32, which approximates the dimensions of the landslides on Black Rapids Glacier. The mass balance, $b_m$, under the debris-covered area was set to zero. The upper surface was iteratively updated according to the difference between mass balance and emergence velocity.
4 RESULTS

4.1 Ice velocity from SAR and ground surveys

Figure 3 shows the surface-parallel velocity field in the ablation area of Black Rapids Glacier in October 1995, obtained with SAR interferometry using ERS tandem data. The pattern and magnitudes of the velocity match those reported by Fatland et al. [2003] for January 1992 and December 1995, as well as survey velocities from 1995 (M. Truffer, personal communication, May 2011). The longitudinal velocity profile (Figure 3, inset) shows that the velocity just above the Loket tributary (km 23) is 50% of that between km 15 and km 20. The longitudinal velocity below the Loket tributary recovers to almost the values between km 15 and km 20. A second slowdown, smaller both spatially and in magnitude than that observed at the Loket tributary, occurs at about km 34, where a minor tributary joins Black Rapids Glacier. A velocity plateau is evident where another minor tributary joins the trunk glacier at km 30.

Five cross-glacier velocity profiles from the ablation area are also shown in Figure 3. The outer inflection points on each profile mark the transition from active to dead ice or moraine. The irregular form of profiles D and E reflects the influence of several minor tributaries and moraines.

The seasonal evolution of surface ice flow in the vicinity of the landslides is shown in Figures 4 and 5. These results were obtained with the speckle tracking method using RADARSAT-1 and ALOS Palsar data. The velocity fields between 2003 and 2007 (Figure 5) are more variable and spatially heterogeneous than the velocity field in 1995 (Figure 3). Monthly velocities across BRG-west and BRG-middle (Figure 4A) are typically higher than across BRG-east, the lowest debris sheet. The highest seasonal velocities are in late May and June 2005 and in May and June 2007. Across chips 1-4 (see Figure 4A for locations of “chips”, or image subsets), which correspond to BRG-west and the upper half of BRG-middle, ice velocities ranged from around 23 cm d⁻¹ between late May and mid-June 2005, to around 19 cm d⁻¹ from mid-June to early July 2005. The upper margin of BRG-west had a velocity of nearly 22 cm d⁻¹ in June 2007, but the velocity decreases to 12 cm d⁻¹ in July of that year. By the end of
the melt seasons in 2003-2007, ice velocities were nearly the same (5-8 cm d\(^{-1}\)) across the area covered by all three landslides.

Time series of surface ice flow rates derived from ground surveying, InSAR and radar speckle tracking (annual pairs) are shown in Figure 6. We calculated representative SAR-derived velocities across the debris sheet in spatially averaged chips and as point measurements for survey station km 29 (Table 2). Figure 6A shows the mean velocity over the upglacier and downglacier chips of the debris sheet (chips 1 and 10, respectively), and the mean of all chips across the entire debris sheet (chips 1-10), averaged for each annual measurement in a particular melt season. Figure 6B shows the reduction in longitudinal velocity gradient with time, towards less compressional flow, from \(-5.7 \times 10^{-3}\) a\(^{-1}\) in 1995 to \(-4.1 \times 10^{-3}\) a\(^{-1}\) in 2004 and only \(-1 \times 10^{-3}\) a\(^{-1}\) in 2007.

Point velocities calculated by InSAR and optical surveying at stations km 20 and km 29 in 1995 are similar (Table 2). The mean 1995 InSAR-derived velocity across chips 1-10 is also similar to the surveyed ice velocity at km 29, which is located about halfway along this reach of the glacier. Despite an acceleration between 1995 and 1997 (Figure 6A), surveyed velocities at km 20 (Figure 6A) were the same in 1995 and 2002, before the earthquake. Similarly, the surveyed velocities at km 29 are about the same in 1995 and 2002. We thus infer that the mean velocity in the vicinity of the landslide debris immediately before the earthquake was approximately equal to the mean InSAR velocity for 1995 (Figure 6A).

The annual surveyed velocities above the landslides (km 11, 14, and 20) show a short-lived, small acceleration (6-14%) in 2003, followed by a deceleration to values slightly less than in 2002 (Figure 6A). Survey data for 2009 show that the glacier was still slowing down at km 14 and km 20, although the exact velocity pattern at km 20 between 2005 and 2009 is unknown, because no survey measurement was made in 2007. We assume that velocity at km 20 is comparable to that measured at km 14, based on the similarity of the two series throughout the period of record.
The survey marker at km 29 was destroyed by one of the 2002 rock avalanches. Because the velocity pattern at km 29 is similar to that at km 20 between 1992 and 2002, we calculated an offset and projected the post-earthquake km 20 data to km 29 for 2007 (dashed continuation of km 29 line in Figure 6A). Using this approach, we estimated what the velocity at km 29 might have been in the absence of the landslides. The continued slowdown at km 29 in the absence of a landslide is supported by numerical modeling (see next section).

SAR speckle tracking measurements downglacier of km 20 reveal spatial and temporal velocity changes that appear too variable to be explained by natural evolution of the glacier, without the effect of the landslides. Notably, the measurements indicate a rapid increase in velocity by 2004. The mean SAR-derived velocity for the entire debris sheet in 2004 is 44% higher than the velocity derived from the 2002 terrestrial surveying at km 29, which is located approximately in the middle of area covered by the debris sheet (Figure 6A, Table 2). The mean velocity in the area of the debris sheet decreases in each subsequent year at a greater rate than at survey sites higher on the glacier. Between 2004 and 2007, the mean velocity of the debris-covered ice decreased 30%, approaching the same velocity as measured by InSAR in 1995. Over the same period, ice velocity at site km 14 decreased by only 20%. In this context, the large velocity variations higher up on the glacier in the previous decades should be kept in mind.

The velocity of the upglacier margin of the debris (chip 1; open circles in Figure 6A) was the same in 1995 and 2004, as measured by SAR. Over the same period, the velocity of the downglacier margin of the debris sheet (chip 10; asterisks in Figure 6A) increased 109%. Between 2005 and 2007, the velocity of chip 10 continued to increase (Figure 6A), while the velocity of the downglacier half of the debris sheet decreased slightly.

The spatial pattern of velocity also changed after the landslides in 2002. In 1995, prior to landslide deposition, the velocity pattern could be characterized as generally decreasing towards the terminus, with significant decreases due to influences from tributary glaciers (Figure 3). Radar speckle tracking from 2004 until 2007, records progressive downglacier increases in ice velocity and a generally more
spatially heterogeneous velocity pattern (Figure 5). The highest recorded annual velocity (~22 cm d\(^{-1}\); 17 May 2005 to 12 May 2006) is at chip 6 at the downglacier margin of BRG-middle.

4.2 Ice velocity from finite element model

In our model control run with no landslide debris (Figure 7), surface velocities decrease slightly over much of the glacier between 2002 and 2007, because the driving stress is reduced, mostly by thinning. The differences are greatest around km 20, which experiences a reduction of velocity of about 9%.

Below km 30, there is little change in velocity over the five-year period, whereas above km 5, the velocity increases slightly due to thickening of the glacier.

Between km 25 and km 32, the main area of interest in this study, flow is compressional in 2002 (-2.9 x 10\(^{-3}\) a\(^{-1}\)) and changes only slightly with time (-2.6 x 10\(^{-3}\) a\(^{-1}\) in 2007) in the control run. Between km 25 and km 28.5, which corresponds to the upglacier half of the area covered by landslide debris, the mean ice velocity without debris is consistently about 3 cm d\(^{-1}\) higher than between km 28.5 and km 32 (Figure 7E).

After adding a debris sheet between km 25 and km 32 in the model, the upglacier end of the debris slows and the downglacier end accelerates, resulting in a switch from compressional to locally extensional flow by 2005 (Figure 7B, F). The velocity gradient between km 25 and km 32 changes from -2.9 x 10\(^{-3}\) a\(^{-1}\) in 2002 (compressional flow) to 2.0 x 10\(^{-3}\) a\(^{-1}\) in 2007 (extensional flow). At km 25, the ice velocity decreases 27% between 2002 and 2007; in contrast at km 32, the velocity increases 56% over the same period. Much of the change in total velocity can be explained by changes in basal motion (Figure 7D). For example, at km 32, the modeled surface velocity in 2002 is ~9.9 cm d\(^{-1}\), with 7.1 cm d\(^{-1}\) coming from basal motion. By 2007 (with landslide debris), total motion increases to ~15.4 cm d\(^{-1}\), with 11.1 cm d\(^{-1}\) due to basal motion and the remainder due to ice creep.

Figure 8 shows the changes in the full-Stokes modeled 2D centerline velocity profile with and without the debris sheet. If the glacier is not covered by a debris sheet, the velocity vectors change little
over five years (compare Figure 8A and 8B), and flow is primarily emergent. With the addition of debris however, the velocity becomes strongly submergent at the upglacier end of the debris sheet (Figure 8C), and slightly less emergent at the downglacier end due to locally extending flow across the debris sheet.

5 DISCUSSION

Post-landslide changes to the surface velocity field of Black Rapids Glacier in the vicinity of the rock avalanche debris sheets include a substantial, but short-lived initial speed-up followed by a gradual slowdown to pre-earthquake values (Figure 6). The velocity changes under the debris are more rapid and of a higher magnitude than changes higher on the glacier, which are discussed below. Superimposed on these trends is a reduction of the longitudinal velocity gradient in the area covered by the debris sheet – the velocity of the downglacier half of the debris sheet increases, while the velocity of the upglacier half decreases or remains constant. Notably, the entire debris-covered part of the glacier trends towards a uniform surface velocity over the period of observation due to a reduction of surface slope of the debris-covered ice.

Gardner and Hewitt [1990] noted a similar change from compressional to extensional flow following deposition of three large landslides on Bualtar Glacier. In the ten months following the landslides, ice near the leading edge of the debris sheet moved four times faster than ice near the trailing edge. In the following 13 months, however, this pattern reversed, with ice near the trailing edge moving twice as fast as that near the leading edge.

The ice surface velocity upglacier from the debris sheets on Black Rapids Glacier increased slightly between 2002 and 2003. We propose three possible explanations for this speed-up. First, it may not represent a change in ice velocity, but rather an offset on the Denali fault during the earthquake. Right-lateral offset near the terminus of Black Rapids Glacier was about 4 m (about 1 cm d\(^{-1}\) equivalent) [Haeussler et al., 2004]. However, elevated velocities were observed over two consecutive seasons (2002-2003 and 2003-2004) at several survey sites (e.g. km 11 and km 14), suggesting that earthquake
displacement is probably not the cause of the increase in speed. Second, a short-lived speed-up may have been caused by an earthquake-generated change in the subglacial plumbing system. Although many authors have examined relationships between ice velocity and water pressure [e.g. Truffer and Iken, 1998; Kavanaugh and Clarke, 2001; Truffer and Harrison, 2006], none to our knowledge has specifically examined a possible relationship between earthquake shaking and till failure by excess pore water pressures. The observed speed-up at km 8 (M. Truffer, personal communication, May 2011), which is not on the Denali fault, casts doubt on this explanation, although the possibility of longitudinal coupling between the Denali fault and the ice at km 8 does not necessarily preclude it. Third, the speed-up may be associated with the decadal-scale oscillations noted earlier. Nolan [2003] reports that velocity oscillations during the quiescent phase of the glacier have a period of about 12 years. The velocity increase in 2002-2004 is smaller and has a shorter onset phase than that of the ~1992-2001 cycle, which itself is smaller and shorter than the first recorded cycle from ~1980 to 1992. The rapid termination and small magnitude of the most recent oscillation may be indicative of a trend towards more temporally constant flow. With only two or possibly three such oscillations, however, this argument is tenuous. Regardless, the speed-up recorded by surveys at km 8, 11, 14, and 20 is small and likely unrelated to the much larger speed-up of the debris-covered ice between km 25 and km 32 identified with the SAR speckle tracking.

Several authors [Gordon et al., 1978; Shulmeister et al., 2009] have suggested that landslide debris may alter glacier flow. The bulk density of rock avalanche debris is probably about 2400 kg m\(^{-3}\) [Vacco et al., 2010], equivalent to about 2.6 m of ice-equivalent for every meter of debris. In the case of thin glaciers, an additional few meters of rock debris may increase the driving stress sufficiently to cause the glacier to speed-up by changing the deformational velocity alone. Black Rapids Glacier, however, is more than 450 m thick in the area of the landslides, and a few meters of rock debris alone is unlikely to alter the velocity significantly. A simple calculation of deformational velocity,
\[ v = \frac{2A}{(n+1)} (\rho g \sin \alpha)^{1/(n+1)} h^{(n+1)}, \] where \( \rho \) is ice density (917 kg m\(^{-3}\)), \( g \) is gravitational acceleration (9.8 m s\(^{-2}\)), \( \alpha \) is the glacier surface slope, and \( h \) is ice thickness (450 m), shows that an increase in ice thickness equivalent to 2 m of rock debris would result in less than a five percent increase in deformational velocity. More important factors are changes in mass balance resulting from strongly reduced ablation beneath the debris, changes in the vertical velocity under the debris resulting from a change to extending flow, and changes in the surface slope of the glacier. Changes in slope result from differential ablation at the upglacier and downglacier margins of the debris sheet and greater emergence velocities at the downglacier end than at the upglacier end. Both of these effects pivot the glacier surface towards the horizontal. In addition, steep ice cliffs develop at the upglacier and downglacier margins of the debris sheet due to the differential ablation. During fieldwork in 2007, the upglacier end of the 2-m-thick BRG-west debris sheet was perched on a pedestal of ice about 15 m high that dipped upglacier. The downglacier end of the BRG-west debris sheet similarly rested on a pedestal about 15 m high, but dipping downglacier. Thus the surface slope at the upglacier end was reversed, whereas the slope at the downglacier end was steeper than in 2002. An increase in ice thickness of 20 m, equivalent to 15 m of unmelted ice and 2 m of rock avalanche debris, results in nearly a 20% increase in velocity.

In our full-Stokes model (Figures 7, 8), the upglacier end of the debris sheet slows while the downglacier end speeds up in response to changes in surface slope and ice thickness. The emergence velocity adjusts to the switch from compressive flow in the ablation area to locally extending flow over a horizontal distance of 7 km. Basal velocity (Figure 7D), which is largely governed by surface slope in the Weertman-style sliding law used here, is reduced at the upglacier end of the debris sheet where a slope reversal occurs due to reduced or no ablation under the debris. Conversely, basal velocity at the downglacier margin of the debris increases due to a steep ice cliff caused by differential ablation. The qualitative match between the model results and the SAR-derived velocities indicates that the modeled scenario is a reasonable representation of the short-term evolution of the surface velocity of a glacier.
covered by landslide debris. A shortcoming of our model is that we do not explicitly consider subglacial hydrology, even though changes in plumbing beneath a glacier, due for example to drainage of a supraglacial lake [Sturm and Cosgrove, 1990], certainly have an effect on surface motion.

The spatial velocity pattern observed with SAR speckle tracking at Black Rapids Glacier (Figure 4A) is more complex than the model would suggest. In spring 2007, the velocity across the BRG-west debris sheet was very high (>20 cm d\(^{-1}\)) while the velocity across the BRG-east debris sheet was much lower (<6 cm d\(^{-1}\)). By early summer of 2007, surface velocities were much more uniform across all three debris sheets (~10 cm d\(^{-1}\)). The short-lived acceleration at the upglacier end of the debris sheets may be related to the introduction of meltwater to the basal plumbing system from drainage of a supraglacial lake. Remote sensing data show that a supraglacial lake forming in spring 2007, dammed by the upglacier margin of the BRG-west debris sheet. The lake had mostly drained by mid-July 2007 (Figure 9). In the 24 June 2007 RADARSAT-1 image, the lake was approximately 140 m wide and 840 m long.

Assuming a constant glacier surface slope of 2\(^{\circ}\) we estimate that the lake contained a minimum of \(~2.9 \times 10^5\) m\(^3\) of water. By 18 July, the lake contained only \(~1.5 \times 10^4\) m\(^3\) of water. The lake is not visible on the 7 May or 31 May SAR images, but the reflectance of the pixels is very low, indicating a surface saturated with meltwater. During fieldwork in summer 2007, we observed the last remnant of the lake (Figure 9D), as well as a moulin about 3 m in diameter at the upglacier margin of the lake. The moulin drained the much larger lake earlier in the melt season.

The highest velocities measured by speckle tracking in 2007 occurred between 31 May and 24 June; the lowest velocities occurred in early autumn. We suggest that slow leakage from the supraglacial lake in May and June 2007 kept water pressure high, leading to low effective pressure, and enhanced basal sliding. In late June, the lake drained through one or more moulins, perhaps creating an efficient tunnel system, with a concomitant reduction in basal sliding and thus surface speed. This ice dynamics scenario is similar to that described for outburst floods at Kennicott Glacier. Several authors [Anderson, S P et al., 2003; Anderson, R S et al., 2005; Bartholomäus et al., 2008] have found that subglacial jökulhlaups
from Hidden Creek Lake have overpressurized the efficient conduit system, forcing water into an inefficient linked cavity system. The resulting high subglacial water pressures cause temporary bed separation and enhanced basal motion. Once the subglacial conduit system evolves to greater efficiency, the augmented basal motion ceases.

Truffer et al. [2005] noted that the low ice velocities measured at km 8 and km 14 on Black Rapids Glacier in 2004, which are some of the lowest in 32 years of record, are likely the result of an efficient water drainage network established at times of high runoff. Meier et al. [1994] described “extra-slowdown” events at Columbia Glacier in Alaska, when, after a speed-up event caused by enhanced input of water to the bed, the glacier slowed, ultimately reaching a velocity that was lower than that before the speed-up. Rapid transfer of a significant volume of water from a supraglacial lake to the glacier bed may have a similar effect at Black Rapids Glacier. Drainage of a supraglacial lake, however, is not a necessary precursor to a spring speed-up – Black Rapids Glacier is temperate-based, thus high water pressures can be expected whenever the basal drainage system is inefficient.

Alternatively, the spring speed-up may be caused by a hydraulic barrier at the bed, instigated by the reversed surface slope at the upglacier end of the rock avalanche debris. If large enough, this surface bulge could change the hydraulic potential gradient, causing a subglacial water pocket to form, which could lead to partial decoupling of the glacier from the bed. A lake drainage as discussed above, could then help drain the water pocket through an efficient pathway.

6 CONCLUSIONS

This study reports the impacts of three concurrent earthquake-triggered landslides on the behavior of Black Rapids Glacier. Observations and numerical modeling suggest that Black Rapids Glacier responded to a landslide-induced shut down of ablation by reducing the local velocity gradient.

Surveying above the landslide debris sheets indicates a small acceleration of the glacier immediately after the earthquake (2002-2004) and a widespread slowdown thereafter. InSAR and satellite radar
speckle tracking (2003-2007) show a large increase in velocity in the area of the landslide debris sheets, especially at the downglacier end of the lowest debris sheet, where velocities nearly doubled. A full-Stokes, numerical ice-flow model produced similar results, suggesting that changes to mass balance, surface slope and a switch from compressional to extensional flow are responsible for the observed changes in surface velocity.

7 ACKNOWLEDGEMENTS

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8 REFERENCES


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Figure Captions

Figure 1 (A). Map of Black Rapids Glacier showing the trace of the Denali Fault (dotted line), the three landslide debris sheets (dark grey shading), and distances from the head of the glacier in 5 km increments [modified from Amundson et al., 2006, reprinted from the Journal of Glaciology with permission of the International Glaciological Society]. Delta River is shown with a wavy pattern. Locations of Fairbanks and Black Rapids Glacier are shown on the inset map with an open circle and star, respectively. (B) Photograph of the BRG-east debris sheet looking west. The BRG-middle debris sheet is in the background. Photo courtesy of Dennis Trabant (US Geological Survey).

Figure 2 (A) Glacier geometry, (B) net balance, and (C) mesh used in the ice dynamics model. Dashed box in (A) represents area shown in (C).

Figure 3 Surface-parallel velocity of Black Rapids Glacier in October 1995 overlain on a geocoded SAR amplitude image. The inset is a profile of ice velocity along the glacier centerline. The black strips are the locations of transverse velocity profiles, shown at the right. The ascending ERS scenes are restricted to the ablation area of the glacier; as a result the velocity map terminates just upstream of profile A at the bend near the equilibrium line. Dashed box outlines region of interest shown in panels in Figure 4. The Loket tributary enters the main trunk of Black Rapids Glacier around km 25, causing a reduction in surface velocity just above the confluence.

Figure 4 Black Rapids Glacier velocity field derived from RADARSAT-1 speckle tracking. (A) 31 May – 24 June 2007. (B) 11 August – 4 September 2007. Survey stations km 20, 26, 29, and 32 are shown in (A), along with chip outlines. Landslide debris sheets are outlined in yellow.

Figure 5 Twenty-four day leapfrog speckle tracking velocity fields for the period 2003-2007. (A) RADARSAT-1 speckle tracking maps in the vicinity of the rock avalanche debris sheets. Rock avalanche debris sheets are outlined with yellow dashed lines in bottom-right panel. (B) Average velocity at each debris sheet based on RADARSAT-1 and ALOS Palsar data.

Figure 6 Magnitude of surface ice velocity derived from ground surveying, InSAR, and speckle tracking. (A) Mean of uppermost and lowermost chips on debris-covered ice (chips 1 and 10). Solid lines record terrestrial surveying data; circles, dots, and asterisks represent SAR-derived velocities. (B) Longitudinal velocity gradient from 1995 to 2007 for the part of the glacier covered by landslide debris. The 1995 data point is an InSAR-derived velocity for October 1995, averaged over the same chips as for the speckle tracking (Figure 3) and converted to an annual equivalent velocity [Rabus and Fatland, 2000].

Figure 7 Modeled glacier velocities calculated using the full-Stokes model. Panels A and B show the horizontal surface velocity for, respectively, the control run and the run with landslide debris from km 25 to km 32. Panels C and D show the basal velocity for the control run and the landslide debris. Open circles in panels E and F show that the mean velocity from km 25 to km 28.5 (upglacier half of debris-covered area); asterisks show the mean velocity from km 28.5 to km 32 (downglacier half of debris-covered area); and the solid line represents the velocity gradient between km 25 and km 32. For clarity, model results are shown only for km 15 to km 40.

Figure 8 Centerline velocity profiles for (A) 2002 prior to landslide, (B) the 2007 control run (no landslide debris), and (C) the 2007 run with landslide debris cover. Arrows indicate velocity directions and represent only a small subset of the velocity vectors in the model. The model was run for the entire
40-km long glacier, as shown in Figure 2; for clarity, only the area covered by, or adjacent to, the landslide debris is shown here.

Figure 9 SAR images showing (A) a melting glacier surface on 31 May 2007, (B) a large lake upglacier of BRG-west on 24 June 2007, and (C) the much smaller, drained lake on 18 July 2007. (D) Photo collage of the lake on 8 July 2007. Glacier flows from left to right in all panels.
Table 1 Details of radar satellite images used in this study.

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Table 2 Surface velocities (cm d⁻¹) measured by ground surveying, InSAR and and radar speckle tracking.

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The InSAR data does not extend to the km 14 survey station – thus only stations km 20 and km29 are listed for 1995. The speckle tracking data have coherence only on the debris-covered ice; therefore we provide only a point measurement at survey station km 29.
Control run (without landslide)

With landslide

A              B

C              D

E              F

Control run (without landslide) With landslide

Surface velocity (cm d$^{-1}$)

Basal velocity (cm d$^{-1}$)

Centreline distance (km)

Surf. vel. (cm d$^{-1}$)

Basal vel. (cm d$^{-1}$)

Year

Surf. vel. (cm d$^{-1}$)

Vel gradient

A              B

C              D

E              F

Control run (without landslide) With landslide

Surface velocity (cm d$^{-1}$)

Basal velocity (cm d$^{-1}$)

Centreline distance (km)

Surf. vel. (cm d$^{-1}$)

Basal vel. (cm d$^{-1}$)

Year

Surf. vel. (cm d$^{-1}$)

Vel gradient

A              B

C              D

E              F